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PROCESSES IN THE STRATOSPHERE BASED ON ROCKET SOUNDING DATA

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ABSTRACT

This paper presents characteristic features of the temperature regime in the stratosphere revealed mainly in analyzing the data of rocket soundings carried out by the USSR during the IGY. These are: types of thermal stratification, features of annual temperature variation, some information about thermal regime in the upper stratosphere during the Arctic polar night and possible causes leading to the formation of temperature inversion.

The atmospheric layer extending from the tropopause to an altitude of 50 km can to a certain extent be called a transitional layer between terrestrial processes and solar processes. The thermal regime of the lower stratosphere is caused primarily by the influence of the underlying surface and the troposphere, whose influence apparently extends up to an altitude of 30-35 km. The thermal regime of the upper stratosphere is primarily due to the influence of short-wave solar radiation and also possibly due to corpuscular radiation. The main portion of the ozonosphere is located in this layer, whose influence on the thermal regime is quite great. /58*

Regular sounding of the atmosphere by meteorological rockets was performed during the IGY in the USSR. Certain conclusions regarding the characteristics of the stratosphere temperature regime and its vertical stratification were drawn from an analysis of the results obtained from the rocket sounding. It was found that the temperature stratification of the stratosphere is not the same for different geographical regions and, in addition, changes as a function of the time of year. According to the works (Ref. 1, 2) all of the forms of stratosphere temperature stratification can be divided into four types which are distinguished by an unusual vertical temperature distribution in the layer between the tropopause and the beginning of temperature inversion.

The type I, called the winter polar type, is characteristic - as its name indicates - for the winter period of polar regions (both the Arctic and the Antarctic). It is characterized by a temperature drop with altitude, with a gradient of 1°/km on the average.

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* Note: Numbers in the margin indicate pagination in the original foreign text.

Type II stratification is characterized by an isotherm, and is most frequently encountered in the transitional seasons both in the polar and in the mean latitudes.

Type III, called the summer type, is characteristic for the Arctic and the Antarctic and for mean latitudes in the summertime. It has a well-defined temperature inversion, with a gradient of about 1°/km.

Thus, these three types of stratification are differentiated according to a seasonal criterion and according to latitudinal zones.

Type IV stratification is tropical. It is encountered in the tropics, is characterized by a high tropopause, and a somewhat more complex configuration: strong inversion above the tropopause (2°/km), and then a layer of small gradients. In addition to the above properties, all four types of stratification are characterized by an increase in the tropopause and stratopause with an increase in the type number, and also by a temperature increase above the tropopause and stratopause for types II - III.

The types of atmospheric temperature stratification are the result of certain processes occurring in the atmosphere (both in the troposphere and in the stratosphere), and therefore can serve as characteristics of these /59
processes. It was found that the winter polar type I is encountered in the moderate zone during the advection of polar masses, and type IV is found in the mean latitudes during the invasion of tropical masses.

A study of the structural characteristics of the stratosphere above the troposphere cyclones and anticyclones, which was performed with the aid of temperature stratification curves, showed that in the summer the temperature field structure of the stratosphere up to an altitude of 30 km is closely related to the baric relief of the troposphere: regions of cold and warmth, respectively, are located in the stratosphere above high and low pressure regions at low levels (500 mb). If this relationship is expressed in percents, it can be shown that this amounts to 90.5% at 14 km; 68.8% at 24 km; 50% at 28 km; and 42% at 30 km. Above 30 km, this relationship is noticeably attenuated, although in certain individual cases it can be traced up to 35-40 km. Above 40 km, as a rule, this relationship is not observed.

It can be seen from an examination of the types of atmospheric temperature stratification that the stratospheric layer being studied is not homogeneous and can be divided into two layers, according to its thermal properties: a lower layer (25-35 km) and an upper layer (35-50 km). The boundary between these layers is the beginning of temperature inversion (isopause). This level, which is one of the very active levels in the atmosphere, separates many phenomena which can be observed above and below it. Its altitude is not constant, and ranges between 30-35 km on the average. The upper and lower layers differ in terms of the daily and annual temperature variation and in terms of several other characteristics, which primarily indicate a difference in the influence of the main thermal

regime factors. Let us examine the temperature field characteristics of these layers in greater detail for polar and moderate latitudes. The thermal regime properties of the polar zone are apparently only slightly related to the radiation regime: if the advent of the polar night (October-February), when the stratosphere is not exposed to direct solar radiation, leads to strong cooling of air in the lower layer, then in the upper layer - where the temperature drop changes by inversion,which remains until the end of the polar night - processes which are completely different from those in the lower layer will undoubtedly have an influence. We shall discuss these processes below.

As is known, the daily temperature variation is extremely small during periods of the polar night and polar day. However, temperature fluctuations both during one day and for several hours can be significant, as can be seen from an analysis of rocket sounding data. These non-periodic temperature changes may possibly be caused by the vertical movement and advection of air masses. The greatest variability of these temperature fluctuations can be observed in the winter: it is almost 1.5-2 times greater in the upper layer than in the lower layer. During the winter, temperature fluctuations during the day at the polar latitudes amount to 2-4° on the average, and in individual cases (for example, when there are anomalous temperature rises) the temperature can change by 20-30°, and sometimes more. During the summer the temperature fluctuations during the day are the smallest during the annual cycle, and amount to $\pm 1°$. Kh. P. Pogosyan and his co-workers (Ref. 3-5) reached similar conclusions for altitudes up to 30 km by analyzing radiosonde measurements.

In contrast to high latitudes, transformation of the air masses is the main process occurring in the moderate zone up to an altitude of 30-35 km. The temperature stratification is caused by the interaction of incoming air masses with the local air. The daily temperature variation in this zone - in contrast to the polar zone where it is barely expressed over a long period of time (polar day and polar night) - is primarily caused by radiation factors and to a certain extent by vertical streams. The amplitude of the daily temperature variation is not great, and /60
amounts to 1.0-1.5° at altitudes of 25-30 km. It amounts to 3-5° at altitudes of 40-50 km. The daily temperature variation, with a maximum around the sunset and a minimum around the sunrise, is frequently disturbed by non-periodic temperature changes related to the advection of masses or to vertical motions.

An analysis of data from rocket launches shows that the magnitude of these non-periodic temperature changes throughout one day can greatly exceed the amplitude of the daily temperature variation. A quantitative determination of the effect of radiation and non-radiation factors on temperature change, for a period of time ranging from sevral hours up to 2-3 days, showed that the effect of non-radiation factors is 10-25 times greater than the radiation factors in the 25-30 km layer; this effect is three times greater in the 35-40 km layer. Thus, although the role of radiation factors increases with an increase in altitude above sea level,

with temperature changes of such brief duration these factors are not fundamental.

A detailed examination of the factors causing temperature changes throughout the day confirms the assumption presented above. Specific calculations for individual cases indicate that vertical heat exchange and advection of air masses play a basic role in the temperature change for a small time interval which is close to 24 hours. Just as in the polar zone, the mean temperature fluctuations during the day in summer are small in this zone, and they barely change with altitude. The greatest mean temperature fluctuations during the day are observed in winter, and reach +4, +5° on the average for the 25-30 km layer. The maximum values of these fluctuations can reach 10-15° and greater in individual cases.

Annual temperature variations are different for different latitudinal zones. At the polar latitudes, the annual temperature variation is well-defined: the curves have the form of a sinusoid which is characteristic for the entire layer. The greatest temperatures are observed in July in the lower layer, and at the end of June - in the upper layer. As one moves away from the poles (with a decrease in latitude) the shift in the greatest temperatures from one layer to another increases. The lowest temperatures in the annual variation also shift in terms of altitude, and this shift is much more sharply defined than in the former case, namely: in the lower layer (up to 30 km) a temperature minimum is observed in December, and in the upper layer - at the beginning of December and in November. In contrast to the shift in the highest temperatures, this phenomenon is attenuated with a latitudinal decrease. Thus, a shift in the highest and lowest temperatures is only observed in the upper layer, and this is apparently due to the diverse nature of the main thermal regime sources in the upper and lower stratosphere.

The temperature amplitude during the annual variation in this zone increases with an increase in latitude, and fluctuates between 24-47°. The nature of this change with altitude is the same both at 80°, and at 70 and 60°, namely: the greatest annual amplitudes are observed in the transitional layer of 30-35 km, reaching 40-45° here. Both above and below this level, the magnitude of the amplitude decreases. In order to explain this phenomenon, one must take the fact into account that the 30-35 km layer is the upper level at which a temperature drop is observed in the winter. Both above and below this level, the temperature increases. In the moderate zone the annual temperature variation has its own characteristic features. In contrast to the polar zone, the temperature curves have the form of a smooth, arched curve. These curves straighten out more and more as the latitude decreases, and in the tropical zone they become almost straight with small bends. In the upper layer, the curve for the annual temperature variation for a latitude of 60° (i.e., at the boundary between the polar and the moderate zone) takes on the form of a sinusoid, which is typical for the polar latitudes. In this layer, the curve for /61
the annual temperature variation for a latitude of 40° (i.e., at the

boundary between the moderate and the tropical zones) takes on a form which is characteristic for curves of the tropical zone. In the lower layer no such sharp division between the curves for 60, 50, and 40° is observed. A subsequent feature of the annual temperature variation in the moderate zone is a shift of the highest temperatures, which is observed only in the upper layer where a temperature maximum is not observed in July, but at the beginning of summer and even at the end of spring. At the present time, no comprehensive explanation of this fact has been found, although many researchers attribute this to the influence of ozone.

In the summer, the structure of the temperature field in the stratosphere and the lower mesosphere at moderate latitudes is fairly simple: the horizontal gradients are small throughout the entire layer (1-2° for 10° latitude). On the other hand, in the winter this zone represents a transitional zone from the tropical latitudes to the polar night. The horizontal temperature gradients at this time are 3-5 times greater than in summer, and the temperature field varies to the greatest extent.

Both in the polar zone and in the moderate zone, the advent of spring warmth is initiated at the higher altitudes and gradually extends below. However, this process takes place much more slowly and not as intensively as it does in the polar zone, where the heating process is completed not only more rapidly but also extends much farther below. Thus, throughout the entire polar day the 25-50 km layer is much warmer at the high latitudes than this same layer is at the mean latitudes, and considerably warmer than in the tropics where there is apparently no such phenomenon by which the layers are heated.

In autumn the layers in both one and the other zone are cooled; this occurs most intensely at the polar latitudes. For two months (September-October) the 25-50 km layer is cooled by 20° or more on the average; in the moderate zone this cooling is 2-2.5 times less. Therefore, beginning in September the entire 25-50 km layer becomes warmer at the moderate latitudes than at the polar latitudes. Thus, the sign change in the pole-mean latitudes temperature gradients occurs twice throughout the year: at the end of April - beginning of May, and at the end of August - beginning of September.

Let us dwell in greater detail on the features of the upper stratosphere thermal regime during the polar night in the Arctic. According to data derived from the rocket sounding over O. Kheys (80° N), the strong temperature inversion in the upper stratosphere is retained until the end of the polar night. This can be seen from data presented in the table, where the local time at which the sounding took place and the height of the Earth's shadow H_T at the zenith, at the moment at which the meteorological rocket was launched, are indicated for each sounding date. Computations show that the radiation balance of these stratospheric layers is negative during the polar night (Ref. 6). /62

TEMPERATURE OF THE AIR IN THE STRATOSPHERE (°K) BASED ON MEASUREMENTS OVER O. KHEYS IN THE WINTER MONTHS OF 1957-1959

H, km	10/22/1957 6 hr 20 min, H_T = 3 km	10/25/1958 11 hr, H_T = 5 km	11/4/1957 10 hr, H_T = 26 km	11/4/1958 15 hr, H_T = 26 km	12/16/1957 23 hr, H_T = 175 km	12/21/1957 11 hr, H_T = 180 km	12/27/1958 11 hr, H_T = 175 km	2/26/1958 6 hr, H_T = 0 km	Summer (Mean)
45	256	-	-	-	-	255	261	-	272
40	232	237	236	237	247	234	242	227	262
35	228	219	227	235	222	226	215	233	249
30	215	210	214	200	208	208	197	223	243
25	211	209	211	208	200	200	193	222	237
20	221	203	221	207	203	203	207	217	236
15	223	-	223	213	213	213	214	215	232
10	222	-	222	213	219	219	214	213	228

Generally speaking, the heating origins of the upper stratosphere can be found in the circulation processes - advection or adiabatic heating during descending motion. In the first case, an assumption would have to be advanced regarding the presence of a winter stream of thermal air at the 30-60 km level from the illuminated portion of the atmosphere, in the region of the Earth's shadow. In the second case, the existence of a similar, predominant winter current would have to be assumed in the higher region (at altitudes of 60-80 km). Subsequent descent to the 40-50 km level would have to be accompanied by adiabatic heating (Ref. 1, 2). If the air descent occurs in the polar mesosphere from a higher level - for example, 100 km - then, in addition to the adiabatic heating, the air descent would have to be accompanied by recombination of atomic oxygen with the liberation of considerable dissociation heat.

We cannot now analyze in greater detail the above dynamic mechanisms by which the atmosphere is heated at high latitudes during the polar night. We can only point out that an estimate of the heat liberated when oxygen is recombined points to very high values.

However, there is a basis for assuming that corpuscular radiation from the sun (Ref. 1, 2) can play a significant role in heating the upper stratosphere during the polar night. As is known, under the influence of the geomagnetic field this radiation can reach the nocturnal side of the earth. In addition, it is even possible that the nocturnal side is preferred:

for example, the aurora polaris apparently appear primarily on the nocturnal side of the earth. The energy of protons in the corpuscular streams bombarding the polar regions of the atmosphere comprises 10^3 $erg/cm^2 \cdot sec$ in the case of a mean magnetic disturbance up to 10^5 $erg/cm^2 \cdot sec$ for a strong disturbance - i.e., from 0.1 to 10% of the solar constant. This energy can serve to considerably increase the temperature of high atmospheric layers. Recent studies by J. Van-Allen and his co-workers have shown that electrons penetrating to the 80 km level are included in the composition of a primary stream of solar corpuscles. Their ionizing action, and consequently thermal action, is manifested indirectly in the atmosphere at lower altitudes, by means of electromagnetic X-ray radiation. Measurements performed on stratosphere balloons have shown (Ref. 7) that this type of ionization extends to 30 km. However, observed temperature inversion commences at the 30 km level. Thus, from this point of view corpuscular radiation of the sun may possibly cause the polar atmosphere to be heated during the polar night.

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